in
"Understanding the Earth".
Cambridge University Press
1992

CHAPTER

3

THE EARTH'S INTERIOR

Don L. Anderson
California Institute of Technology

Introduction

Seismologists have developed many methods for studying the structure and composition of the Earth's interior. Tens of thousands of earthquakes occur every year and each one sends out seismic waves in all directions. The bigger events send waves into the deep interior and these are used to infer the three-dimensional structure of our planet. Large earthquakes also generate surface waves which travel repeatedly around the surface and constructively interfere to generate the free oscillations, or normal modes, of the Earth. The velocities of seismic waves depend on composition, mineralogy, temperature and pressure and are useful as tools for studying these parameters. The new science of seismic tomography (meaning to 'slice the Earth') provides images which are useful in mapping convection.

While we can sample the crust of the Earth by a variety of techniques, the rest of the interior is largely inaccessible. Geophysical data, such as the mass and moment of inertia of the Earth and the velocities of seismic waves, are not adequate unambiguously to infer the chemistry of the mantle and core. We know from the distribution of elements in the Solar System which elements are likely to be most important in the Earth. We also obtain valuable clues by studying meteorites, the properties of the other planets in the Solar System, and the Sun (see also Chapters 1 and 2).

Nebular condensation and planet formation

It is commonly assumed that the Sun and planets formed more-or-less contemporaneously from a common mass of interstellar dust and gas, known as the solar nebula. There is a close similarity in the relative abundances of the condensable elements in the atmosphere of the Sun, in chondritic meteorites and, probably, in the Earth. To a first approximation we can assume that the planets incorporated the condensable elements in the proportions observed in the Sun and the chondrites. On the other hand, the differences in the mean densities of the planets, corrected for differences in pressure, show that they cannot all be composed of materials having exactly the same composition. Variations in iron content and in the oxidation state of iron can cause large density variations among the terrestrial planets, whereas, the giant, or Jovian planets, must contain much larger proportions of low atomic weight elements than Mercury, Venus, Earth, the Moon and Mars.

The equilibrium assemblages of solid compounds that exist in a system of solar composition depend on temperature and pressure, and, therefore, on location and time relative to nebular condensation (Figure 3.1). At a nominal nebular pressure of 10^{-4} bar the material would be a vapour at temperatures greater than ~ 1800 K. The first solids to condense at lower tem-

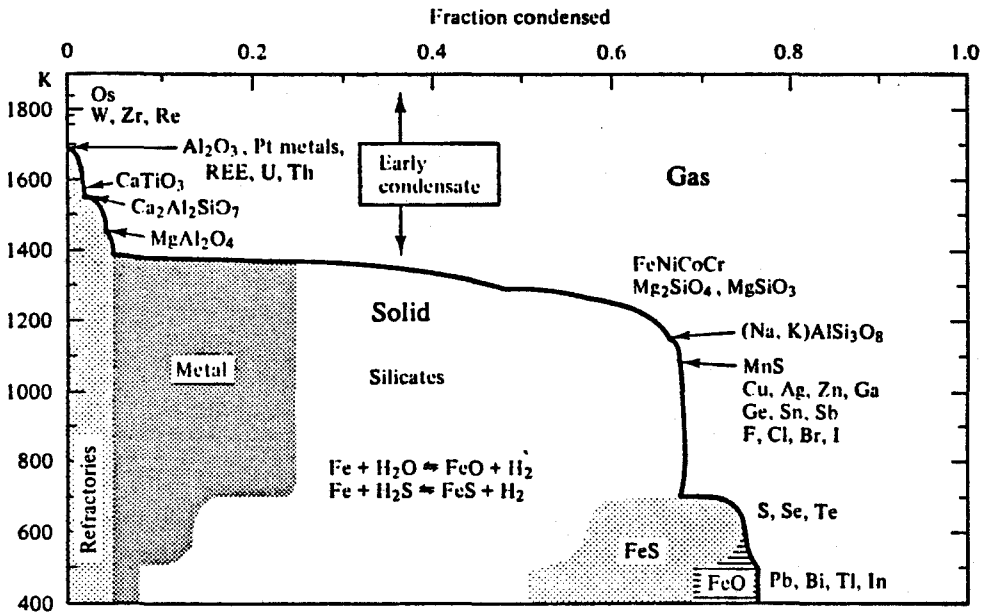


Figure 3.1 **Condensation of solar gas at 10^{-4} bar represented in terms of the fraction condensed against absolute temperature (see also Chapter 2). REE = rare earth elements.** □

perature or higher pressure are the refractory metals such as W, Re, Ir and Os. Below about 1700 K refractory oxides of more abundant elements such as Al, Ca, Mg and Ti condense, and metallic Fe condenses near 1400 K. Below ~ 1150 K, Na and K form feldspars and a portion of the Fe is stable in olivine and pyroxene with the proportion increasing with a further decrease in temperature. FeS forms below ~ 750 K. Hydrated silicates condense below ~ 300 K. Differences in planetary composition may thus depend on the location of the planet, the location and width of its feeding zone and the effects of other planets in sweeping up material or perturbing the orbits of planetesimals, the small bits of pre-planetary material. In general, one would expect planets closer to the Sun to be more refractory rich than the outer planets. On the other hand, if the final stages of accretion involved coalescence of large objects with solar orbits of different eccentricities then there may be little correspondence between bulk chemistry and the position of the terrestrial planets.

There are several ways in which interactions between gaseous nebula and solid condensate particles might have controlled the composition of the planets. At one extreme, all or most of the material joining a planet may have equilibrated in a relatively narrow range of temperatures peculiar to that planet. Alternatively, one mineral after another condenses as required to maintain thermodynamic equilibrium in the nebula, and immediately accretes into a planet, or into a planetesimal which eventually joins a planet. At some point the process is interrupted by dissipation of the uncondensed nebular remnants. Differences in the mean compositions of the planets would result when the gaseous nebula, with any remaining uncondensed elements, was removed. If temperatures declined outward in the nebula at the time when condensation ended, this could account qualitatively for the density differences in the planets. The planets formed may thus have been layered from the outset, having the highest-temperature condensates at their centres, and successively lower-temperature condensates closer to their surfaces. Such planets would contain substantial metallic cores mainly as a result of accretion rather than subsequent interior melting and differentiation. This is termed **heterogeneous accretion** and is in contrast to the situation where all the solids have completely condensed before serious planet building starts: **homogeneous accretion.**

Modern theories of planetary accretion are of hierarchical nature. There is assumed to be a range of sizes in the pre-planetary body population and some large objects were involved in the formation of each planet. For example, an Earth-sized body may have coalesced from one Mars-sized and ten Moon-sized objects and numerous smaller bodies during its accre-

tionary stage. In fact, current theories for the origin of the Moon have the Moon resulting from the splash between the proto-Earth and a Mars-sized impactor (details in Chapter 2). Large impacts tend to melt, or even vaporise, the colliding objects and to throw a large amount of material out into space. The Earth may have been extensively melted several times before the major part of the impacting population was used up and incorporated into the surviving planets. The present internal structure of the Earth and its chemical stratification may be a result of these early processes occurring during accretion. Where gas pressures are high, for example near a growing planet, or immediately after a giant impact, solids, melts and gases can coexist. At low nebular pressures most gases condense directly to a solid as indicated in Figure 3.1.

Cosmic element abundances and the composition of the Earth

The planets and the Sun formed from a diffuse nebula of gas and dust that collapsed sometime just after 4600 million years (Ma) ago, the age of the Earth (see Chapter 2). The terrestrial or 'rocky' planets and the meteorites can be thought of as rocky debris from this event, most of the mass having entered the giant planets and the Sun. Since the deep interior of the Earth is inaccessible we use data from the Sun and meteorites, as well as geophysical constraints, to infer what the Earth may be made of.

The term volatile elements is often used to refer to elements such as H, He, C and S and compounds containing these elements which are gaseous until very low temperatures and which are not easily accreted into planets. In contrast, compounds containing the refractory elements*, Mg, Si, Ca, Al, Fe etc., condense as solids at relatively high temperatures (Figure 3.1) and are easily incorporated into growing planets. Elements such as K, Na, Rb and Pb are of intermediate volatility and apparently may be partially lost during the high-temperature accretional stages of a planet.

Of course, the Sun is mainly H, He and C, which are very volatile elements, but it also contains small amounts of Mg, Si, Fe, Ca, Al and O, the main constituents of the meteorites and rocky planets. After the Sun formed, most of the more volatile elements were swept out of the vicinity of the inner Solar System, probably at an early period when the solar wind was much more intense. The terrestrial planets are therefore composed primarily of the more refrac tory non-volatile or condensable elements, whereas the volatile elements are the primary building blocks of the Sun, the giant planets, and some of the satellites in the outer part of the Solar System. Comets are also composed primarily of volatile elements and compounds. Through solar spectra the Sun provides us with estimates of the ratios of the refractory elements such as Fe/Si, Mg/Si, Al/Si and so on. These ratios are likely to be close to those in the Earthas a whole.

Table 3.1. *Solar and meteoritic (CI carbonaceous chondrites) atomic abundances of some elements (relative to Si)*

	Sun*	CI†
Na	0.067	0.0574
Mg	1.089	1.074
Al	0.0837	0.0849
Si	1	1
P	0.0049	0.0010
S	0.242	0.0515
K	0.0039	0.00377
Ca	0.082	0.0611
Ti	0.0049	0.0024
Fe	1.270	0.900
Ni	0.0465	0.0493

* Breneman, H. H. & Stone, E. C. (1985) *Astrophysics, J.*, 299, L57.

† Anders, E. & Grevesse, N. (1989) *Geochim. et Cosmochim. Acta*, 53, 197.

Table 3.1 gives the composition of the Sun, as derived from spectroscopic studies and analysis of particles emitted by the Sun. Also given is the composition of a class of meteorites called the carbonaceous chondrites, Type 1 (CI). These meteorites have compositions close to the refractory portion of the Sun (i.e. the refractory elements occur in solar ratios) and also contain a large fraction of volatile elements (such as C and S).

Various attempts have been made to estimate the composition of the solar nebula by combining solar and chondritic chemical composition data. These are the so-called 'cosmic abundances' which are the reference composition for the planets and other meteorites. Materials which have nearly cosmic abundances of a group of elements are called 'primitive' or 'undifferentiated' or 'chondritic'. The CI carbonaceous chondrites are the most solar-like in their composi-

* Refractory elements are those that condense at fairly high temperatures, say 1400 K or more. We define volatile elements as those that condense at lower temperatures, say 1200 K or less.

tions and have higher abundances of volatile elements than other types of meteorites. They are therefore called 'primitive' even though there is reason to believe they have experienced low-temperature alteration, possibly on the surface of a small Solar System object, or inside of a comet. Other meteorites appear to have been processed at higher temperatures, either in space or in small bodies such as asteroids, satellites or near the surface of small planets.

With some idea of the relative importance of the various chemical elements in our Solar System we can construct crude models of the Earth and other terrestrial planets. The Earth has a heavy core which is 32.5 per cent by mass of the planet. A glance at the cosmic abundance table (Table 3.1) shows that iron (Fe) is the only abundant heavy element so it is reasonable to assume that most of the core is Fe. It probably also contains about 6 per cent Ni. On the other hand, the core is a little less dense than pure Fe or Fe–Ni. There is a variety of elements, the siderophile (iron-loving) elements (see Figure 3.2), which may be mixed with Fe in the core, but these are unlikely to reduce its density. However, oxygen is an abundant element which dissolves in molten Fe at high pressure. FeO may also be an important component of the lower mantle. We therefore assume that the core is:

$$Fe + FeO = \text{'}Fe_2O\text{'}.$$

This has about the right density and there is enough 'Fe_2O' in cosmic abundances to make a core about the right size. We write Fe_2O in quotes since there is no guarantee that there is such a compound. There is probably some sulphur in the core as well, but sulphur is a volatile element and may not have been accreted by the Earth in cosmic abundances (see also discussion in Chapters 2 and 4). But there is a possibility that chalcophile (sulphur-loving) elements are concentrated in the core.

The major rock-forming refractory elements are Mg, Si, Al, Ca, Fe and O and their abundances in the silicate Earth are estimated in Table 3.2. In minerals these occur as MgO, SiO_2, Al_2O_3, CaO and FeO, Fe_2O_3 or Fe_3O_4. For example, the minerals enstatite (pyroxene) and forsterite (olivine) can be written:

$$MgO + SiO_2 \rightleftharpoons MgSiO_3 \text{ (enstatite)}$$

and

$$2\,MgO + SiO_2 \rightleftharpoons Mg_2SiO_4 \text{ (forsterite)}.$$

We assume that these elements are fully oxidised in the crust and mantle (with Fe as FeO). Fe can substitute in these minerals for Mg, and the FeO-rich end-members are ferrosilite and fayalite respectively. On this basis, Table 3.2 (column 1) shows the oxide composition of 'primitive' mantle (mantle plus crust). Note that the remaining cosmic Fe that is not needed in the core is enough to make 15 per cent FeO for the mantle. This is much greater than the proportion found in most rocks and magmas which occur in the upper mantle, so this suggests that upper mantle rocks may not be typical of the whole mantle (see upper mantle peridotite composition, column 2 in Table 3.2). Some ancient lavas, called komatiites, do have a very high FeO content (~14 per cent – see Table 3.2, column 3).

Table 3.2. *Estimates of the composition of the silicate Earth (i.e. mantle plus crust)*

	(1)	(2)	(3)
SiO_2	45.0	44.2	44.9
Al_2O_3	3.2	2.0	3.1
FeO	15.7	8.3	13.5
MgO	32.7	42.3	33.0
CaO	3.4	2.1	3.8

(1) Computed from solar abundances, with 32.2 wt % core (1 mole Fe + 1 mole FeO = Fe_2O).
(2) Average peridotite from kimberlite pipe.
(3) Komatiite melt.

Another way to estimate the composition of primitive mantle uses rocks and magmas which are exposed at the surface and which were brought up from great depth by volcanic eruptions. Most basalts come from the mantle, particularly those at ocean ridges, oceanic islands and volcanic arcs. There are also large basaltic lava flows on land, called continental flood basalts, which originate in the mantle. Basalt is the most abundant material from the mantle but it probably is only a partial melt so that a large amount of more refractory material is left behind. Rocks called peridotites, harzburgites and lherzolites are composed primarily of the refractory crystals olivine and orthopyroxene and have only a small amount of calcium, aluminium, sodium and titanium, elements which are important in basalts. Some peridotites may represent the material left behind after basalt extraction. The 'bulk silicate Earth' also includes the crust so this must be mixed back in when estimating the composition of primitive mantle. The crust is only a small fraction of the Earth but it contains a large fraction of some of the so-called 'incompatible' or 'lithophile' (rock-loving) elements.

The relative portions of basalt and peridotite in the Earth are unknown but we can use chondritic or solar ratios based on the refractory elements to decide how much of each is required. Table 3.3 gives an estimate

Table 3.3. *Terrestrial abundance table*

	Mantle and crust	Relative to CI
Li	2.1 ppm	0.87
Na	2040 ppm	0.26 v
Mg	20.52 %	1.46
Al	2.02 %	1.57
Si	22.40 %	1.44
P	57 ppm	0.05 s,v
S	48 ppm	0.0025 s,v
K	151 ppm	0.17 v
Ca	2.20 %	1.58
Ti	1225 ppm	1.86
Fe	6.1 %	0.22 s
Ni	1961 ppm	0.13 s
Rb	0.39 ppm	0.11 v
Sr	16.2 ppm	1.42
Th	0.0765 ppm	1.50
U	0.0196 ppm	1.40

s = These elements may occur primarily in the core.
v = volatile, these elements are depleted in the Earth compared with CI meteorites or the Sun (relative to the refractory elements).

of the composition of the Earth based on these considerations and an extended version with more elements appears as Figure 3.2.

The refractory elements (Mg, Al, Si, Ca, Ti, Th, U ...) all occur at a level of about 1.5 times chondritic (CI meteorites) in the mantle–crust system. This is the level expected if these elements have been excluded from the core and occur only in the silicate part of the Earth. The volatile elements (Na, P, S, K, Rb ...) are depleted in the mantle and crust, relative to chondrites and the Sun, and the elements which are depleted the most are the most volatile or the most siderophile. This depletion probably occurred during the accretion of the Earth, and during core formation. The Fe in the Earth is primarily in the core along with other siderophile elements such as Ni, Co, Re, Os ..., and also, possibly, P and S. The mantle–crust system is extremely deficient in siderophiles. Having developed broad constraints on the chemistry of the Earth's interior the next step is to introduce geophysical data that allow us to construct an amplified picture.

Seismology and the Earth's interior

The Earth's interior is illuminated by seismic rays radiating outward in all directions from the numerous earthquakes that occur in tectonically active regions of the world. The wavefronts developed at right angles to these ray paths are refracted (bent) and reflected by discontinuities and gradients in material properties in the Earth's interior and are recorded at seismic stations on the Earth's surface. From analysis of such waves seismologists have divided the Earth into a crust, mantle and core and into numerous smaller subdivisions such as the upper mantle, transition region, lower mantle, outer core and inner core. These will be discussed later. Box 3.1 discusses the math-

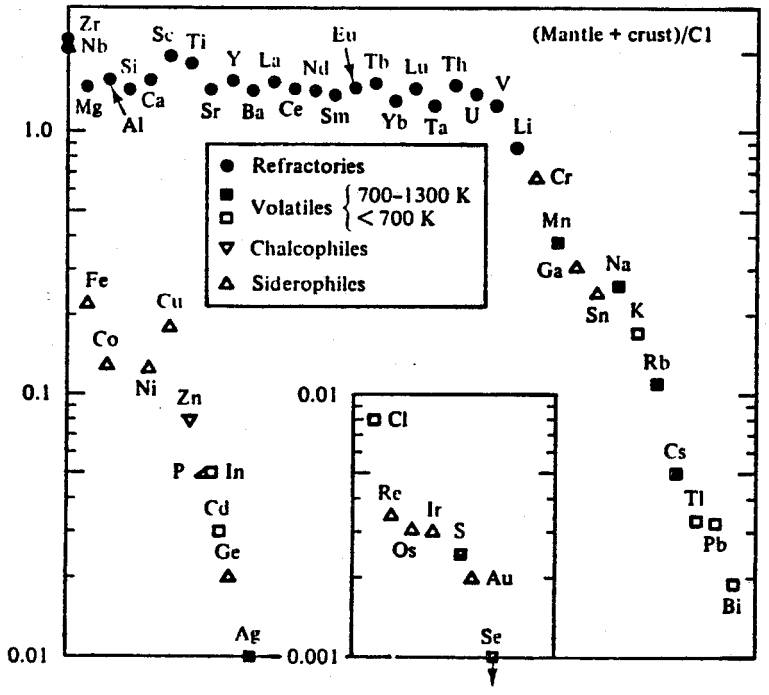


Figure 3.2 Chemical composition of the silicate Earth (mantle plus crust), sometimes called the bulk silicate earth (BSE) or Primitive Mantle relative to the compositions of CI carbonaceous chondrite meteorites. □

BOX 3.1

Seismic wave velocities, elastic moduli and density

An isotropic solid exhibits both a **compressibility** and a **rigidity**. The former describes the resistance of the solid to a change in volume due to a change in confining pressure. The rigidity expresses the ability to withstand a shape change, with no change in volume, due to an imposed shear stress. Liquids and gases are relatively compressible and have no rigidity.

Sound waves in fluids travel with the compressional velocity, V_P:

$$V_P = \sqrt{K_S/\rho}\,, \tag{1}$$

where K_S is the adiabatic **bulk modulus**, the reciprocal of the compressibility, and ρ is the density. In a solid the compressional velocity is:

$$V_P = \sqrt{\frac{K_S + (4/3)G}{\rho}}\,, \tag{2}$$

where G is the **rigidity modulus**.

The shear velocity is:

$$V_S = \sqrt{\frac{G}{\rho}}\,. \tag{3}$$

The first wave to arrive from an earthquake travels at velocity V_P and is compressional; it is often called the P- or primary-wave. The particles in such a wave exhibit a push–pull motion, just as do sound waves in the air (Figure 3.3). The shear, secondary or S-wave, travels much more slowly, although it usually has a larger amplitude because of the shearing nature of motion along earthquake faults. The motion associated with the shear wave is a side-to-side shaking motion. The particle motion, in fact, is perpendicular to the direction of wave propagation (Figure 3.3).

Seismic compressional waves and most sound waves depend on the adiabatic bulk modulus, K_S. It is called 'adiabatic' because there is little time during the passage of the wave for heat to flow. The compressed parts of the wave warm up the solid but the dilatational part of the wave soon arrives and cools it off. In a static experiment a solid can be compressed very slowly and one can wait until the temperature of the solid equilibrates with the surroundings. In this case the compression of the solid is controlled by the isothermal bulk modulus, K_T. The two are related by a well-known thermodynamic identity:

$$K_S = K_T(1 + \alpha\gamma T) \tag{4}$$

where α is the coefficient of thermal expansion and γ is the Grüneisen parameter, a small number usually between 1 and 2. The coefficient of thermal expansion is typically between 20 and 30×10^{-6} K^{-1}. For a temperature, T, of 1000 K and with $\gamma = 1$ and $\alpha = 20 \times 10^{-6}$ K^{-1}:

$$K_S/K_T = 1.02. \tag{5}$$

ematical details of seismic wave velocities, elastic moduli and density, while some of the seismic wave paths which are used are shown in Figure 3.4.

The nomenclature of Figure 3.4 is quite simple. The primary (compressional) and (shear) secondary waves in the mantle are P-waves and S-waves. Compressional waves in the outer core are K and in the inner core are I. Waves reflected at the boundary of the outer core include the letter c in their designation (e.g. PcP, ScS). Waves reflecting off the inner core boundary are PKiKP, SKiKS and so on. A shear wave through the inner core is called J but evidence for its existence is indirect because no shear waves occur in the outer core. Waves can reflect off the surface of the Earth many times (e.g. Figure 3.4: SS, SSS, PP ...). They also bounce around in the outer core (PKKP, PKKKP, ...).

The travel times of these various waves tell us the depths to **seismic discontinuities** (important physical and chemical boundaries in the Earth) and the variation of seismic velocity with depth and from place to place. The amplitudes of these waves tell us about the size of the seismic event, the reflection coefficients of seismic discontinuities and the attenuation of seismic waves. In general, the velocities of seismic waves increase with depth but in tectonic and volcanic regions there is often a region in the shallow mantle where particularly the shear velocity decreases with depth, due to high temperature gradients near the Earth's surface and the onset of partial melting (Figure 3.5). Pressure compresses the rocks and minerals of the

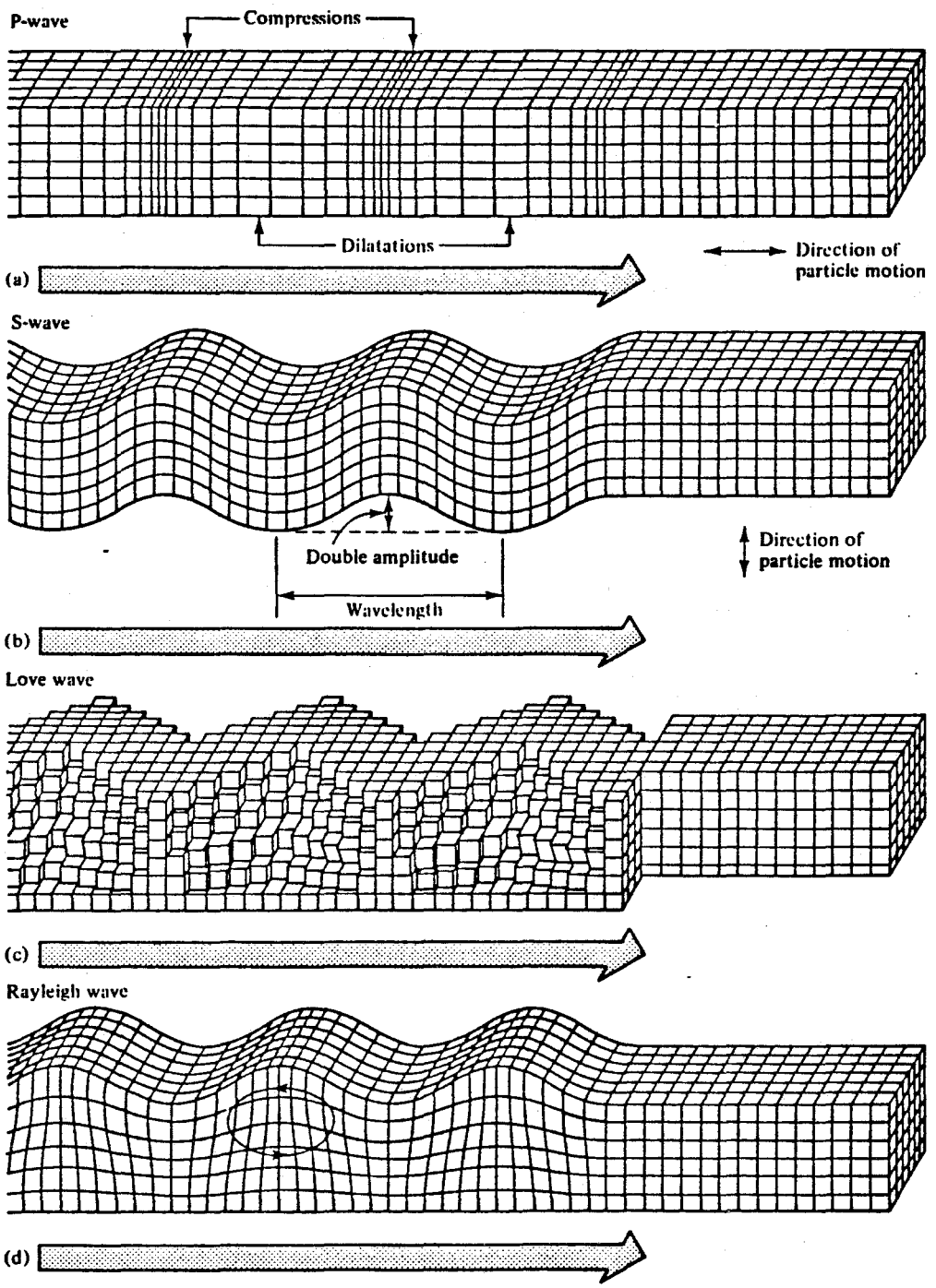


Figure 3.3 The forms of the motion and ground deformation due to (a) P-waves, (b) S-waves, (c) Love waves and (d) Rayleigh waves. The particle motion of P and S-waves is transmitted through the Earth, so these are known as body waves, whereas the motion due to Love and Rayleigh waves diminishes with depth, so these are known as surface waves. □

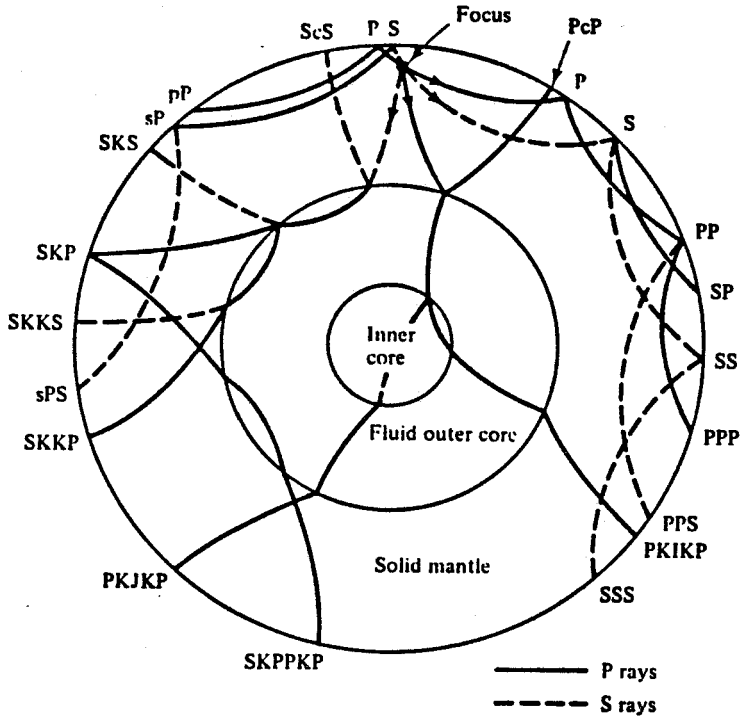


Figure 3.4 Representative seismic wave paths in the Earth all arising from the same focus. P, K and I are compressional waves; S and J are shear waves (see text for further details). □

mantle and causes the density and seismic velocities to increase. Temperature has the opposite effect, and partial melting also decreases density and seismic velocity. In the upper part of the mantle the temperature gradient is high and the tendency for velocity to decrease with temperature is comparable to or greater than the tendency of pressure to increase velocity. At greater depth, pressure is the dominant influence.

The temperature gradient is also high across major chemical interfaces such as the **core–mantle boundary** (CMB). The reason is that heat cannot be convected across such boundaries and a steep temperature gradient – across a 200–300 km thick zone (discussed in more detail later) – is established in order to conduct the heat away from the interior. The shear velocity may also decrease at the base of the mantle, a result of the high thermal gradient. P-waves are somewhat less sensitive to temperature and more sensitive to pressure than S-waves.

Seismic velocities vary with depth, and from place to place at the same depth. For example, stable shield areas not only have thick crusts they also have high velocities in the underlying mantle, a combination of cold temperatures and high velocity refractory minerals. Volcanic regions and regions of young lithosphere, such as near oceanic ridges, have very low seismic velocities, primarily the result of the velocity reduction due to the presence of magma in the shallow mantle.

Figure 3.5 shows the variation of shear velocity with depth in a stable shield area, in an old oceanic area and in a tectonically active area. Note the large variation of velocity above 200 km depth. The seismic lithosphere, or lid, is thick (~ 150 km) under shields and much thinner under the other regions. The near-surface high-velocity regions may correspond to the minimum thickness of the plates that are involved in plate tectonics (see also Chapter 8). The velocities under the lid in tectonic and oceanic regions are so low that no combination of the abundant upper mantle minerals can account for them unless the temperature is so high that partial melting occurs. The presence of a melt phase is implied down to a depth of at least 300 km in these regions. The relatively high velocities under shield areas between about 150 and 400 km have been used by some to argue that continental roots extend this deep and that the continental plate extends deep into the mantle. The alternate point of view is that shield mantle is normal and it is the presence of an enhanced melt phase under oceans and tectonic areas that makes them differ from normal colder, or average temperature, mantle.

The lateral variation of seismic velocity decreases markedly below 200 km (Figure 3.5) and again below 400 km. In fact, the high-resolution body wave (P and

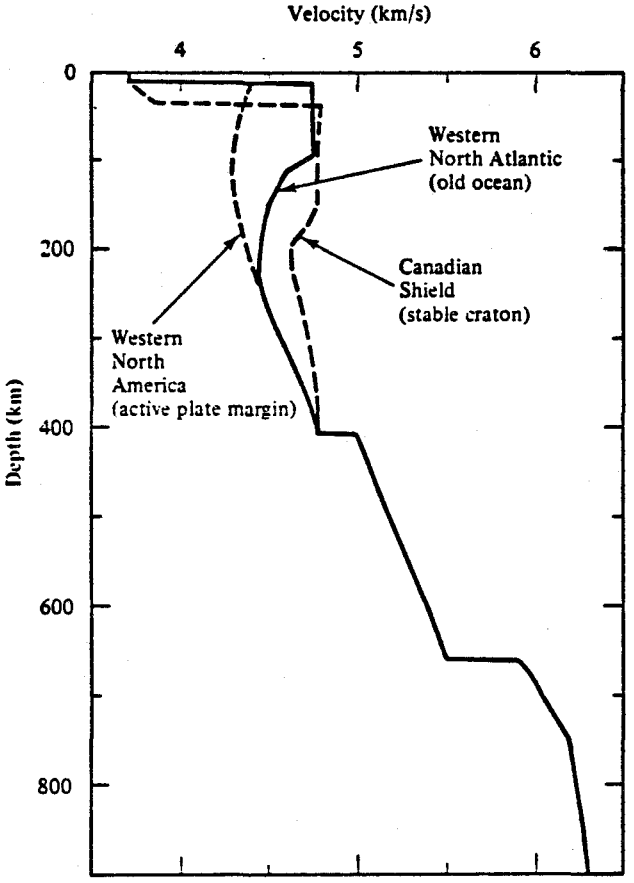


Figure 3.5 **Seismic shear velocity vs. depth in three different tectonic provinces as determined by Helmberger and co-workers at Caltech by comparing theoretical seismograms with observed seismic records. The theoretical mantle model is adjusted until there is good agreement with the data. The travel times of seismic waves at different distances from a seismic source are also used to construct velocity–depth functions.** □

S-waves travelling through the Earth) studies of Professor Helmberger and his colleagues (Caltech) show little if any lateral variation below 400 km in the regions studied. In other studies there are slight variations below this depth. However, the new science of seismic tomography has resolved slight variations down to the core–mantle boundary; these studies will be discussed later.

There are large and abrupt increases in seismic velocity at 5–40 km depth, also near 400 and 650 km depth (Figures 3.5 and 3.6). These are associated with changes in mineralogy as, for example, at the shallowest discontinuity, the crust–mantle boundary or **Mohorovicic discontinuity** (Moho). A change in mineralogy can be due to a change in chemistry, or a change in temperature or pressure with no change in chemistry. For example, the common minerals in the continental crust are quartz and feldspar, a result of the high SiO_2, Al_2O_3, Na_2O and K_2O content of the crust. The main stable minerals at the top of the mantle are olivine and orthopyroxene, a consequence of the higher MgO and lower SiO_2 etc. content. Because of this change in mineralogy there is a large change in density and seismic velocity at the Moho.

A change in chemistry can be more subtle. The large jump in seismic velocity at 400 km is primarily due to the effect of increasing pressure which causes the collapse of olivine and orthopyroxene to denser minerals with the same chemistry (i.e. **phase changes**). At temperatures of 1400–1500 °C these phase changes occur with increasing pressure equivalent to 400 km depth. The discontinuity may therefore be an equilibrium phase boundary between low and high density phases (for details, see Chapter 4). However, it has been suggested by some that the region between 400 and 650 km depth may be richer in a basaltic fraction, and therefore poorer in olivine, than the mantle above 400 km. If there is such a change in chemistry at this boundary, an increase in garnet and clinopyroxene for example, it would be hard to detect because of the large phase change effect. Another large seismic discontinuity occurs at about 650 km depth. It is also primarily due to a change in mineralogy and may also be accompanied by a chemical change. The high velocity gradient region just below the discontinuity (Figure 3.6) may represent a spread-out phase change.

BOX 3.2

Determination of density in the Earth

Zones within the Earth are considered to be homogeneous if there are no chemical variations or phase changes. The change of density, ρ, with radius, r, in any such region is given by:

$$\frac{d\rho}{dr} = \left(\frac{\partial \rho}{\partial P}\right)_T \frac{dP}{dr} + \left(\frac{\partial \rho}{\partial T}\right)_P \frac{dT}{dr}, \quad (6)$$

where the subscripts refer to the parameter (pressure, P, or temperature, T) being held constant. For a homogeneous self-compressed region in which temperature increases because of this compression (i.e. under an **adiabatic temperature gradient**):

$$\frac{dP}{dr} = -g\rho, \text{ where :} \quad (7)$$

$$g = \frac{GM(r)}{r^2}. \quad (8)$$

g is the acceleration due to gravity, G is the gravitational constant and $M(r)$ is the mass inside radius r.

In a convecting mantle the mean temperature gradient, away from thermal boundary layers, is close to adiabatic, denoted by the subscript S.

$$\frac{dT}{dP} = \left(\frac{\partial T}{\partial P}\right)_S = \frac{T\alpha}{\rho Cp}, \quad (9)$$

where Cp is the specific heat at constant pressure and α is the volume coefficient of thermal expansion (cf. Box 3.1) defined as:

$$\alpha = -\frac{1}{\rho}\left(\frac{\partial \rho}{\partial T}\right)_P. \quad (10)$$

It is therefore convenient to write the temperature gradient in the Earth as:

$$\frac{dT}{dr} = \frac{T\alpha}{\rho Cp}\frac{dP}{dr} - \tau, \quad (11)$$

where τ is the departure from the adiabatic gradient (required by the isobaric dT/dr in equation 6).

Adiabatic compression of a material is given by the adiabatic bulk modulus, K_S:

$$K_S = \left(\frac{\rho \partial P}{\partial \rho}\right)_S, \quad (12)$$

where S means constant entropy.

Seismic waves are also adiabatic, and hence we can use (from equations 2 and 3, Box 3.1):

$$V_P^2 - (4/3)\, V_S^2 = K_s/\rho = (\partial P/\partial \rho)_S = \phi \quad (13)$$

to calculate the variation of density with radius in a homogeneous region of the Earth for which we have seismic data. Combining equation 6 with equations 7 and 13, and equations 10 and 11, for a homogeneous adiabatic self-compressed region, we have:

$$\frac{d\rho}{dr} = -g\rho/\phi + \alpha\rho\tau. \quad (14)$$

This is known as the **Williamson–Adams equation.**

Much of the upper mantle and transition region are not homogeneous or close to adiabatic because of large temperature gradients, partial melting and phase changes. In such regions, the term $\alpha\rho\tau$ can be used to correct for non-adiabatic gradients provided there is no chemical change or melting, and, of course, provided α and τ are well known (which is often not the case). Most of the lower mantle, however, appears to be close to homogeneous and adiabatic, so that the equation holds well. The Williamson–Adams equation also cannot be used in parts of the Earth where the chemistry is variable, and here we resort to producing Earth density models by trial-and-error that are constrained by seismic travel times, free oscillation data, and other properties such as the Earth's mass and moment of inertia. The simplified results shown in Figure 3.6 were produced using a combination of these techniques.

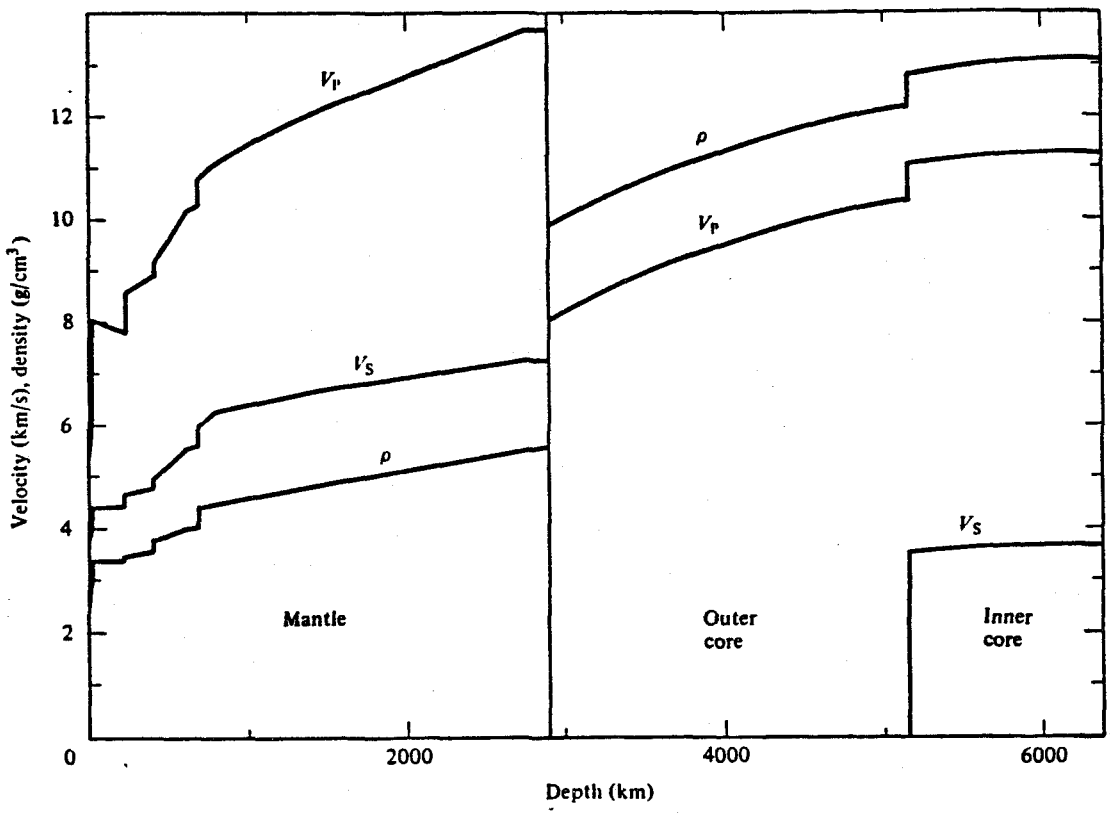


Figure 3.6 Seismic velocities V_p and V_s throughout the Earth and density (π) deduced as explained in Box 3.2. □

The density and seismic velocity throughout the Earth are summarised in Figure 3.6 and a summary of the way in which we derive density data from observed quantities is given in Box 3.2. The boundary between the mantle and the core (CMB) occurs at a depth of about 2900 km. There is a marked decrease in compressional velocity, V_p, and an increase in density as we go from the mantle to the core; moreover, shear waves are absent. These observations are consistent with a molten iron-rich core. The subsequent sections of this chapter amplify details of the structure, composition and mineralogy of the Earth, building on the outline introduced above using Figures 3.5 and 3.6.

Properties of the Earth's layers: some further details

At the centre of the Earth is the solid **inner core**. It is about the size of the Moon but has approximately the density of pure iron or an iron–nickel alloy. It is generally thought to have frozen out of the liquid outer core. The outer core is molten but is probably close to the freezing point. The inner core is therefore suspended in a relatively low-viscosity fluid and is fairly isolated from the rest of the Earth.

The **outer core** extends from the boundary of the inner core (5155 km depth) to the core–mantle boundary which is at a depth near 2900 km, about half of the radius of the Earth (Figure 3.6). The outer core does not transmit shear waves and is therefore a fluid.

The adiabatic bulk modulus, K_s of the top of the core is about the same as that at the base of the mantle but since there is no rigidity in the outer core ($G = 0$), the value of V_p actually drops when crossing from the mantle to the core (Box 3.1, Equation 2). The density of the core is about twice the density of the mantle and this also contributes to the velocity drop. Seismic waves entering the core from the mantle therefore refract downwards (e.g. at B and E, Figure 3.7). They are eventually refracted upwards by the increase of seismic velocity with depth but there is a gap at the Earth's surface in which no direct waves through the top of the core arrive. This is called a 'shadow zone'

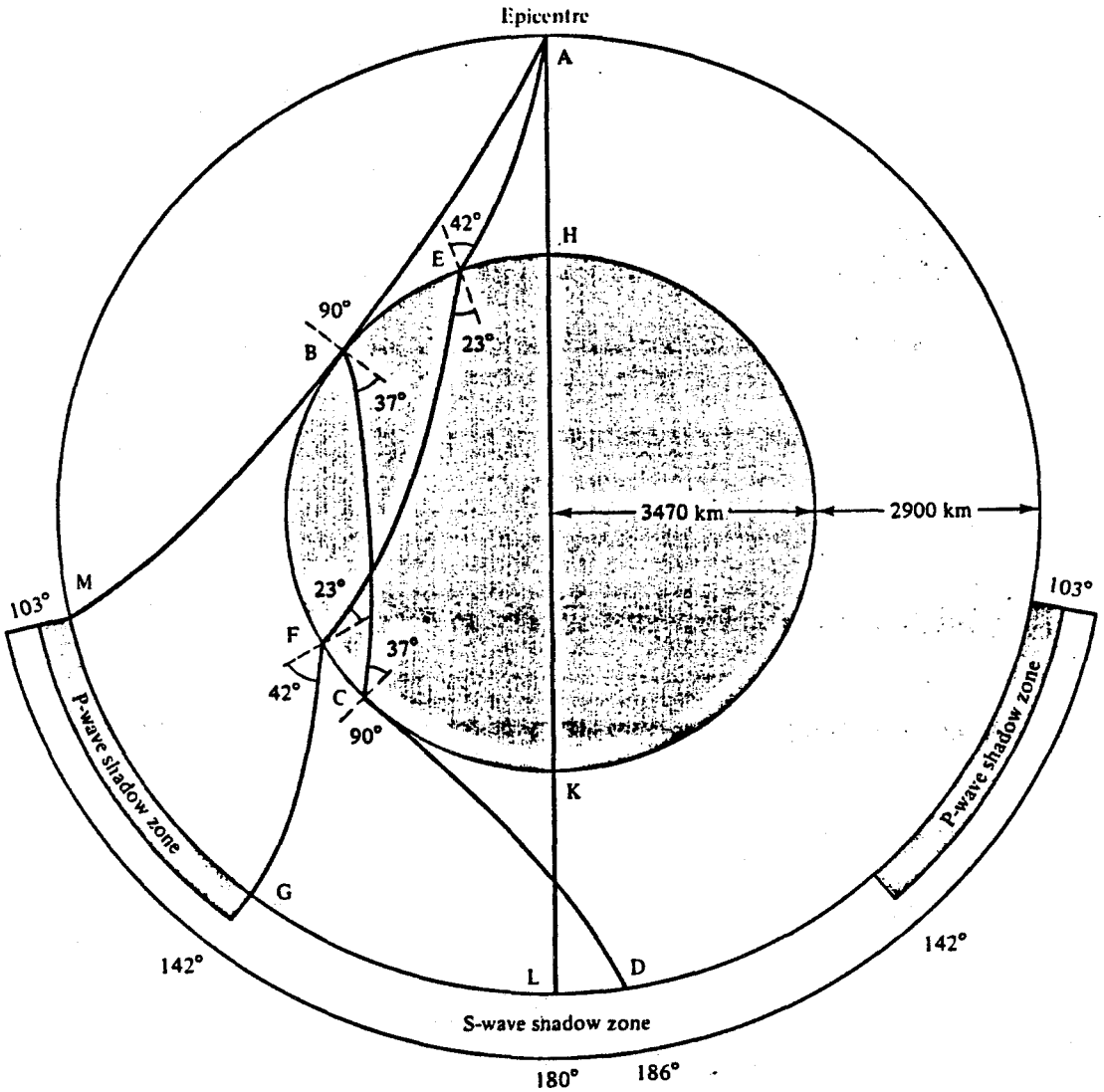


Figure 3.7 The origin of the P-wave shadow zone due to the outer core. A mantle wave path AM just grazes the core and a path only minimally steeper than AB causes the ray to be strongly refracted into the core and out again at C, arriving at D. Waves reaching the core between B and E arrive at the surface between D and G, but as the initial path (e.g. AH) becomes steeper than AE further arrivals at the surface move from G to L. Thus no waves arrive between M and G – the shadow zone between 103° and 142° around the Earth from the epicentre (known as the epicentral angle). This area does, however, receive some seismic energy deflected strongly from the inner core (see Figure 3.4, path PKIKP) but the strength of these arrivals is nowhere near so strong as between 142° and 180°. □

(Figure 3.7) and its size is the primary evidence of the radius of the core.

Various theoretical attempts have been made to estimate the viscosity of the outer core and these generally conclude that it is not much more viscous than water, and may even be more than twenty orders of magnitude less viscous than the mantle. It therefore convects readily and cannot support large internal temperature or chemical gradients. It is probably the most homogeneous part of the Earth but this does not rule out light layers at the top and dense layers at the bottom where material may have settled out. In fact, the inner core may have formed from iron particles settling through the outer core. The core has probably cooled over time and, as it cools, crystals of pure iron may form from the iron–oxygen (and/or sulphur)

mixture which has a lower freezing temperature. Production of these crystals will induce a form of chemical convection and this may help stir the core and provide some of the energy to drive the terrestrial 'dynamo' and generate the Earth's magnetic field. The other observable feature about the outer core is that it transmits seismic waves with very little attenuation (i.e. loss of energy). Once a seismic wave gets into the core it can bounce around for a long time. In contrast, the inner core attenuates seismic waves very rapidly.

Seismic waves reflect well off the core–mantle boundary (CMB), good evidence that the boundary is sharp and exhibits a large change of velocity and density over a fraction of a seismic wavelength. There may be long wavelength bumps on the CMB, of the order of 0.5 to 10 km, but this is still uncertain. However, topographic coupling is one way in which stresses may be transmitted across the CMB. For example, motions in the core could affect the rotation of the Earth by applying pressure to the bumps and valleys. The bumps themselves would be maintained by convection in the high viscosity mantle where relatively highly viscous and rigid, yet lower density material floats on the high density fluid below.

There is an irregular solid layer just above the CMB which has a thickness of 200 to 300 km. This is called the D" layer, a leftover from the nomenclature of Keith Bullen of Australia who proposed a letter designated for each region of the Earth. This layer is variable in both velocity and thickness and in some places is separated from the overlying mantle by a sharp but small discontinuity. In some places the shear velocity gradient in this region is negative, i.e. the velocity *decreases* with depth. These features are best explained by the existence of a chemically distinct layer at the base of the mantle and this layer is also a **thermal boundary layer**. The core is convecting and is losing heat to the overlying cooler mantle (see Chapter 4 for temperatures). However, since little or no material is crossing the boundary this heat must be lost by conduction. There is therefore a region of high thermal gradient just above the CMB which may extend for several hundred kilometres. A high thermal gradient usually means low seismic velocity and density gradients. In a chemically uniform mantle the material in such a thermal boundary layer would become buoyant and the layer would become unstable and rise. Thus it is likely that D" is a chemically distinct layer that is intrinsically much denser, so that the material may be trapped. Nevertheless, it is believed that some sustained mantle plumes may originate in this layer. However, there is no clear evidence that such plumes rise to the surface of the Earth, or even into the upper mantle.

The CMB represents the largest density contrast in the Earth. It is therefore a natural collection point for any light material leaving the core or dense material settling out of the mantle. Thus D" may be a remnant of the original differentiation of Earth, containing material intermediate in density between the core and the bulk of the mantle. Candidate materials of high density are stishovite, a high-pressure form of SiO_2, $(Mg,Fe)SiO_3$ (perovskite) and (Mg,Fe)O (magnesiowüstite). If D" represents early condensing refractory material, or even deeply subducted oceanic lithosphere, it would also be rich in CaO, Al_2O_3 and TiO_2. D" is only about 2 to 3 per cent of the mass of the Earth.

The **lower mantle** is the largest subdivision of the Earth. It extends from the top of D" (about 2740 km depth) to the major seismic discontinuity at a depth near 650 km. The lower mantle is 49.2 per cent by mass, of the Earth and 72.9 per cent of the mantle plus crust. The predominant mineral in the lower mantle is $(Mg,Fe)SiO_3$ in the dense **perovskite structure**, a high pressure form of the chemically identical lower density pyroxene (see Box 3.3 for details of mantle minerals). The second most abundant mineral is probably (Mg,Fe)O in the NaCl or rock-salt structure. Other possible minerals are Al_2O_3 (corundum), different perovskite-structure minerals containing CaO and Al_2O_3, and stishovite. Seismic velocities increase gradually with depth throughout most of the lower mantle and there is no evidence for discontinuities or changes in chemistry. It is thought therefore that the lower mantle is relatively uniform in composition and that the density and seismic velocities increase with depth primarily due to compression – thus the Williamson-Adams equation applies, cf. Box 3.2. Some lateral variations occur, however, primarily because of temperature differences.

The next important feature of the mantle is the **650 km discontinuity** (sometimes called the 670 km discontinuity, but it may vary in depth from place to place by 50 to 100 km). This represents a change in mineralogy between intermediate density minerals, such as γ-spinel and majorite, and high density phases such as perovskites and magnesiowüstite (see Box 3.3 and the next section of this chapter). It may be simply an equilibrium phase boundary, or the two sides of the boundary may also differ chemically in, for example, the contents of FeO, CaO, Al_2O_3 or SiO_2. Phase changes in mantle minerals are generally spread out over at least tens of kilometres depth (see Figure 3.8). But the 650 km discontinuity at 22–24 GPa pressure is a good reflector of seismic energy and this requires that its surface is sharp to about 4 km. It is also close

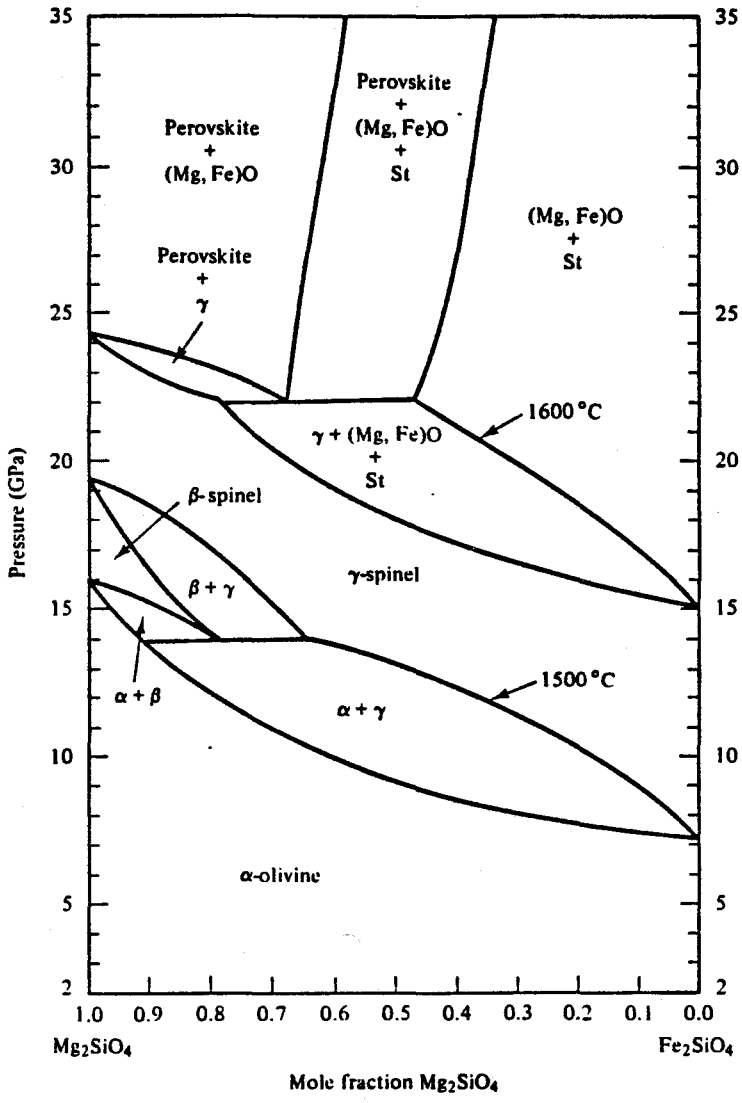


Figure 3.8 Phase diagram of the olivine system $(Mg,Fe)_2SiO_4$, showing the stability fields of the various stable phases, St–stishovite (SiO_2). The α phase is olivine, and both the β and γ phases are forms of spinel. Notice how both the olivine to spinel and spinel to perovskite etc. phase changes occur at greater depths in Mg-rich than in Fe-rock systems. These two sets of phase changes are believed to be important, respectively, at the 400 and 650 km discontinuities. □

to the maximum depth at which earthquakes occur. These features are sometimes taken as evidence that this is a chemical boundary but these arguments are not conclusive. In some Earth models there is a high seismic velocity gradient for 100 km below the discontinuity, and this suggests that the solid–solid phase changes do occur over a broad depth interval even though the upper surface of this zone is a sharp discontinuity.

The region of the mantle between the 650 km discontinuity and another mantle discontinuity at 400 km is known as the **transition region**, being the transition between the upper and lower mantle. The **400 km discontinuity** at 13–15 GPa pressure is due, in part, to the transformation of magnesium-rich olivine to β-spinel (Figure 3.8) and orthopyroxene to a garnet-structured mineral called **majorite**. However, the jump in seismic velocity at 400 km is much less than the jump predicted for these phase changes (see Figure 3.10 and related discussion in the next section) so there must be other minerals near this depth that do not transform. Garnet and clinopyroxene, the main minerals in eclogite, are candidates. **Eclogite** is the high pressure form of basalt and may be more abundant in the mantle than generally thought. The mineralogy of the transition region is therefore thought to include β- and γ-spinel, garnet and majorite. Clinopyroxene is stable at the top part of the transition region but it eventually collapses to a garnet-like phase, probably at about 500 km depth. Indeed, some seismic models have a discontinuity, or at least a high-velocity gradient region, near 500 km.

In the **upper mantle**, down to about 400 km depth, the seismic velocities are consistent with a mixture of

BOX 3.3

Mantle minerals

Atoms and elements (Table 3.3) are the basic building blocks of minerals. Minerals are the basic building blocks of rocks, both in the crust and mantle. The most abundant minerals, in turn, may be considered as composed of the oxides of the elements (Table 3.2). The important upper mantle minerals are olivine, orthopyroxene, clinopyroxene and garnet. These form solid solutions between relatively simple end-members:

forsterite	+	fayalite	=	olivine,
Mg_2SiO_4		Fe_2SiO_4		$(Mg,Fe)_2SiO_4$
enstatite	+	ferrosilite	=	orthopyroxene,
$MgSiO_3$		$FeSiO_3$		$(Mg,Fe)SiO_3$
diopside	+	jadeite	=	clinopyroxene,
$CaMgSi_2O_6$		$NaAlSi_2O_6$		$(Ca,Mg,Na,Al)Si_2O_6$
pyrope	+	almandine	=	garnet
$Mg_3Al_2Si_3O_{12}$		$Fe_3Al_2Si_3O_{12}$		$(Mg,Fe)_3Al_2Si_3O_{12}$

Note that diopside and pyrope can be expressed as $CaO + MgO + 2SiO_3$ and $3MgO + Al_2O_3 + 3SiO_2$ respectively. The above reactions are oversimplified but, nevertheless, by comparing the elements in the above minerals with the abundances in Table 3.3 we see that the most abundant elements are all accounted for. Estimated abundances of major minerals in the mantle lie in the following ranges (weight per cent): olivine 37–51, orthopyroxene 26–34, clinopyroxene 12–17, garnet 10–14.

The above minerals are typical of upper mantle assemblages. At high pressure these minerals convert to denser forms with different crystal structures, a process known as a solid–solid phase change (see Chapter 4 for definitions). For example, diamond and graphite are two solid phases of carbon; diamond is the high pressure and low temperature phase. Natural diamonds were made at depth in the mantle and then were brought rapidly to the Earth's surface in kimberlite eruptions. The important mantle minerals also undergo these pressure-induced phase changes. The mineral forsterite (olivine), Mg_2SiO_4, for example, converts to a dense cubic mineral having a structure similar to normal spinel, Al_2MgO_4, at a depth of about 400 km (13–15 GPa, Figure 3.8). This is referred to as the olivine–spinel phase change. There are two spinel-like forms of $(Mg,Fe)_2SiO_4$, the β-phase structure is about 7.5 per cent denser than olivine (the α-phase) and the γ-phase is about 10 per cent denser. The first phase change (α to β) is probably responsible for the seismic discontinuity in the mantle at a depth of 400 km; at greater depths (pressure) β-spinel converts to γ-spinel and then a further sharp density change to the perovskite structure (see below) occurs at 650 km.

The effects of temperature and pressure on the crystal structure of $MgSiO_3$ are shown in Figure 3.9. Note that in this illustration the chemical composition is constant; the phase boundaries move if impurities such as Al_2O_3 and FeO are introduced. Orthopyroxene, $(Mg,Fe)SiO_3$, is also unstable below about 400 km depth. It recrystallises to a garnet-like phase and can be thought of as dissolving in garnet at high pressure. Note the similarity in the formula of orthopyroxene and garnet, when written in the following form:

orthopyroxene	garnet
$(Mg,Fe)_4Si_4O_{12}$	$(Mg,Fe)_3Al_2Si_3O_{12}$

The garnet form of orthopyroxene is called majorite. Clinopyroxene undergoes a similar transformation at somewhat greater pressure. This mineral is not as well studied at high pressure and it may require the presence of garnet in order to transform to a garnet-like structure. At a depth of 500 km in the mantle all of the familiar low-pressure minerals have disappeared, to be replaced by β- and γ-phase spinels and a complex garnet (majorite) solid solution (see Table 3.4).

At still higher pressure the spinel form of olivine decomposes to an ultradense assemblage containing $(Mg,Fe)SiO_3$ in the perovskite (named for $CaTiO_3$) structure and (Mg,Fe)O, known as magnesiowüstite, in the rock salt (NaCl) structure. The garnet and majorite structure also collapse to perovskite-like minerals (Figure 3.9). These phases are probably stable throughout the whole lower mantle where there are fewer changes and lateral variations in velocity are very small; this is because variations in temperature and pressure occur without changes in mineralogy.

3.3 continued

Table 3.4. *Mineralogy of the mantle*

Depth (km)	Olivine	Orthopyroxene
0	α-phase $^{VI}(Mg,Fe)_2{}^{IV}SiO_4$	orthopyroxene $^{VI}(Mg,Fe)^{IV}SiO_3$
400	β-phase $^{VI}(Mg,Fe)_2{}^{V}SiO_4$	majorite $^{VIII}(Mg,Fe)_2{}^{VI}[(Mg,Fe)Si]^{IV}Si_3O_{12}$
500	γ-phase $^{VI}(Mg,Fe)_2{}^{IV}SiO_4$	
650	perovskite + magnesiowüstite $^{VIII-XIII}(Mg,Fe)^{VI}SiO_3$ + $^{VI}(Mg,Fe)O$	perovskite $^{VIII-XI}(Mg,Fe)^{VI}SiO_3$

In summary, Table 3.4 shows the formulae and coordination numbers (roman numerals) of the cations of the various phases of the two most important mantle minerals, olivine and orthopyroxene. The coordination number is the number of cations of a given type which occur as nearest neighbours to each oxygen atom. For example, in normal silicate minerals at the Earth's surface, there are four oxygens around each silicon ion and six oxygens around each magnesium or iron ion. Coordination number is conserved across some phase boundaries but increases with pressure or decreases with temperature across other boundaries. An increase of coordination is generally associated with an increase of density and seismic velocity and this is a particularly important effect at the 650 km discontinuity.

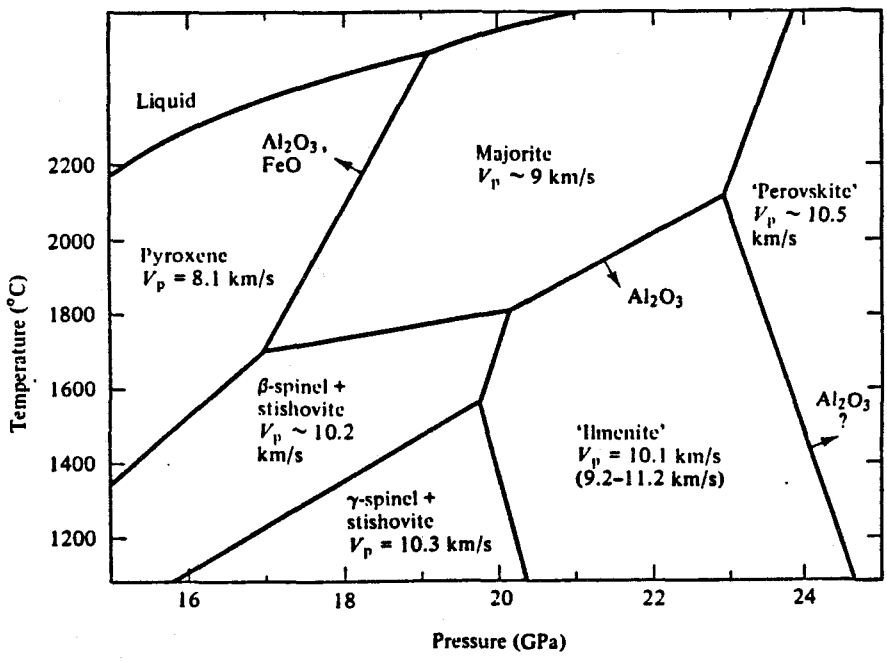


Figure 3.9 **Phase relations in pure $MgSiO_3$, showing the effect of temperature on the phase boundaries. Also given are estimates of the compressional velocity at zero pressure.**

olivine and orthopyroxene (peridotite), or of garnet and clinopyroxene (eclogite), or a mixture of peridotite and eclogite (so-called fertile peridotite). The deepest samples, from kimberlite pipes, come from as deep as 200 km and these are primarily peridotite although eclogites are not uncommon. In the upper mantle it takes a relatively small change of temperature or pressure to change the stable mineral assemblage. Lateral variations in seismic velocity are primarily due to the solid–liquid phase changes caused by such fluctuations in temperature and/or composition. When eclogite melts it turns into a basaltic magma (for details, see Chapters 4 and 5). Basalt is the predominant material that reaches the surface from the interior of the Earth. A variety of evidence suggests that basalts separate from their immediate parent at depths shallower than about 100 km, but the ultimate source region may be much deeper, perhaps in the transition

region. Buoyant upwelling in the mantle is probably responsible for much of the lateral heterogeneity of the upper mantle which is being mapped by seismologists using the methods of seismic tomography (see later).

Seismic velocities of mantle minerals

Seismology is used in several entirely different ways in studying the Earth's interior. It is used as a structural tool and, in this mode, as we have seen, has led to the discovery of the Moho, the outer core and the inner core, and several discontinuities in the mantle. The velocities of seismic waves are sensitive to chemistry and mineralogy and, as will by now be increasingly clear, seismology is also useful as a petrological probe.

Crustal rocks are composed of relatively low density minerals and yet seismic velocities are nevertheless low compared with mantle rocks. The velocities are low (Box 3.1, Equation 2) because the large interatomic spacing has a greater effect on K_S, the adiabatic bulk modulus, than it does on density. The rigidity, G, also generally increases with density, unless the density increase is due to an increased FeO content, or a high concentration of other high atomic weight ions.

Table 3.5 gives some of the properties of common crustal and upper mantle minerals. Note that typical crustal minerals (the upper half of the table) have densities less than about 3 g/cm^3 and compressional velocities less than about 6.5 km/s. The rock types including these minerals have similar properties; for example, quartz and feldspar-bearing granites have low velocities and densities. Basalt has intermediate seismic velocities because it is made up of minerals which, together, have intermediate velocities (clinopyroxene, feldspar). At high pressure basalt converts to eclogite and this is composed of garnet and

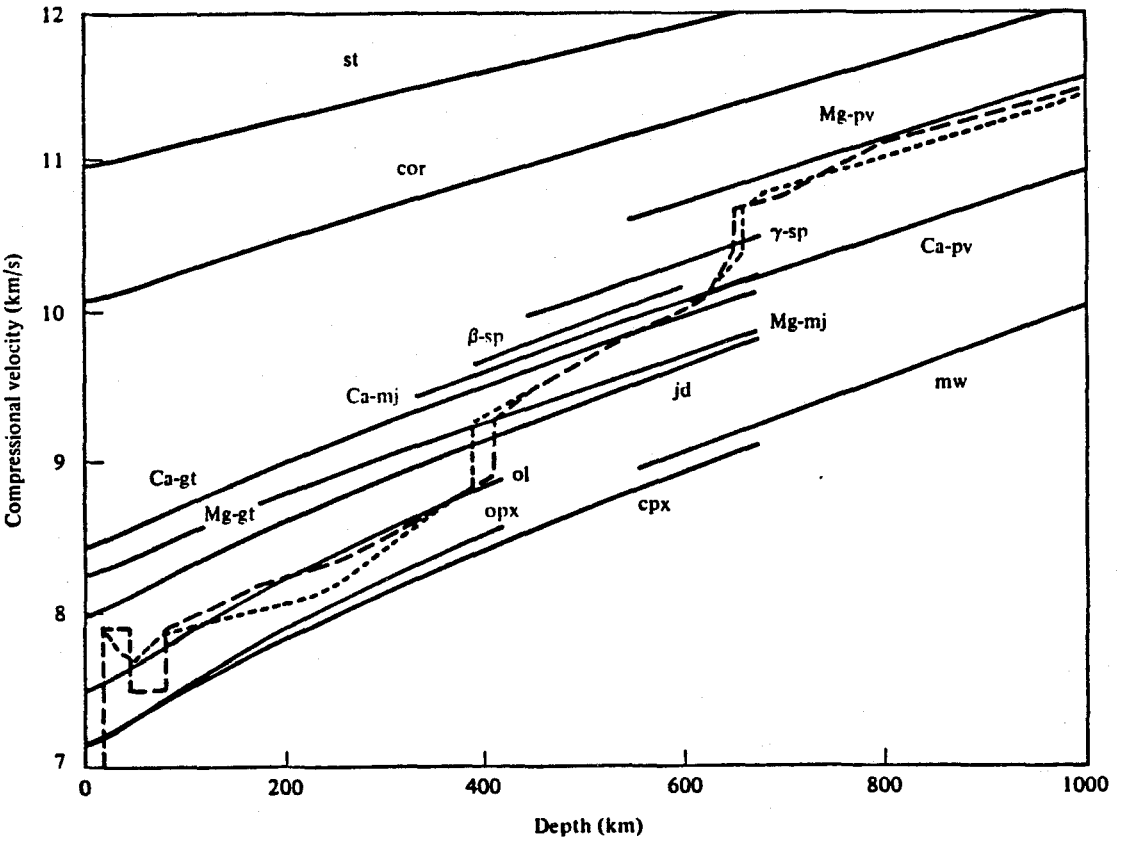


Figure 3.10 Calculated compressional velocities plotted against depth for a variety of mantle minerals. The dashed lines are typical seismic profiles. The calculations are done along at 1400 °C adiabat. (st = stishovite, cor = corundum, pv = perovskite structure, mw = magnesiowüstite, sp = spinel structure, mj = majorite, gt = garnet, jd = jadeite, ol = olivine, opx = orthopyroxene, cpx = clinopyroxene.) Ca- and Mg-rich end-member composites are plotted separately where these strongly influence the compressional wave velocity. □

Table 3.5. *Properties of crustal and upper mantle minerals at typical depths for these minerals (see also Figure 3.10)*

Mineral	Density (g/cm³)	V_p (km/s)	V_s (km/s)
Quartz	2.65	6.05	4.09
K-feldspar	2.57	5.88	3.05
Plagioclase	2.64	6.30	3.44
Mica	2.8	5.6	2.9
Amphibole	3.2	7.0	3.8
Clinopyroxene	3.3	7.80	4.51
Orthopyroxene	3.35	7.84	4.73
Olivine	3.3	8.4	4.9
Garnet	3.56	8.96	5.05

clinopyroxene which, together, yield relatively high densities and seismic velocities. This is one reason why the extent of the eclogite component in the upper mantle is difficult to resolve, geophysically, from that of the olivine-rich peridotite. Equations 2 and 3, Box 3.1, can be used to convert the parameters in Table 3.5 to elastic moduli, e.g. K_S and G.

Figure 3.10 shows the compressional wave velocities in various minerals as a function of pressure or depth, calculated along an adiabat assuming a 1400 °C surface temperature. The dashed lines show a range of typical mantle compressional velocity profiles for several regions. Note that the upper mantle, shallower than 400 km, has velocities close to olivine but ortho- and clinopyroxene can also be present as long as there is enough garnet to counteract the low velocities.

The minerals which are stable below 400 km (Table 3.4) are β-spinel and majorite, the high-pressure forms of olivine and orthopyroxene respectively, and clinopyroxene and garnet. Below about 500 km the stable minerals are a garnet-form of clinopyroxene, majorite, garnet and γ-spinel. Below about 750 km the stable minerals are various forms of perovskite ($MgSiO_3$, $CaSiO_3$ ...) and magnesiowüstite (Mg,Fe)O. There is also some FeO in the perovskites. Stishovite and corundum may also be stable under lower mantle conditions, depending on the average composition.

In order to refine estimates of mantle mineralogy we need to improve our knowledge of mineral seismic velocities and their variation with temperature and pressure. Mineral physics is that branch of geophysics which concerns itself with the measurements and extrapolation of physical properties under extremes of temperature and pressure. One of the most important goals is the determination of accurate 'Equations of State' which will allow the computation of physical properties under the range of conditions found in planetary interiors.

Seismic tomography

Lateral heterogeneity of the Earth's interior is studied by seismic tomography. Earthquakes are used as seismic sources and seismic stations on the Earth's surface serve as receivers. The idea of tomography can be illustrated very simply. The inside of a cave can be studied by a group of people all using flashlights as a source of illumination and their eyes as receivers. By combining their information they can draw three-dimensional maps of the cave. Seismic tomography works in much the same way. Computers replace the mental processing and convert the arrival times of seismic waves to velocity anomalies. Both surface waves (see Figure 3.3) and body waves can be used in tomography. After a large earthquake the surface waves travelling around the world constructively interfere and set up standing wave patterns called normal modes or **free oscillations**. The properties of these waves, their periods and spatial patterns, can be used to infer radial and lateral velocity structure; the longer the period of the standing wave the more effectively it 'sees' to a greater depth in the Earth because of the way in which the amplitude of the disturbance decreases with depth (Figures 3.3c and d). However, some of the normal modes are also due to constructive interference of body waves which travel through the interior and reflect off interfaces.

Figure 3.11 shows maps at various depths of the shear wave velocity determined from surface waves and long-period shear waves by Toshiro Tanimoto. Figure 3.12 shows cross-sections of the upper mantle derived from the surface wave tomographic results of Henri-Claude Nataf, Ichiro Nakanishi and myself. It turns out that there is a remarkable correlation between upper mantle seismic velocities and tectonics, at least down to a depth of 300 km. Active tectonic and oceanic regions have very low shear velocities, in many cases so low that partial melting is implied. In particular, extensional regions (ocean ridges, rifts, back-arc basins) tend to have very low velocities whereas stable continental regions such as shields and platforms have high velocities. Below 300 km, convergence regions tend to have high velocities, evidence for cold subducted slabs.

There are several reasons why seismic velocities vary from place to place at the same depth. The most important are isobaric phase changes, including partial melting, and changes in mineralogy due to chemical changes. Upper mantle minerals are also anisotropic and tend to be aligned by flow processes in the mantle. Changes in mineral orientation therefore can also cause lateral changes in velocity. The total range of

shear velocity variation is about 10 per cent above 100 km, 7 per cent between 100 and 220 km, 4 per cent between 220 and 400 km, 3 per cent between 400 and 670 km and 1 per cent in the middle of the lower mantle. The variation picks up again in D", being about 2 per cent in that region.

Tectonic and young oceanic regions have high mantle temperatures and exhibit volcanism, evidence for melts in the mantle. In the course of continental drift, continents tend to move away from hot areas and come to rest over cold areas of the mantle (see also Chapters 8 and 10). In the process they override cold oceanic lithosphere. This may explain why continents generally have high seismic velocities between 200 and 400 km depth compared to tectonic regions (Figure 3.11a). However, below 400 km convergent bounda-

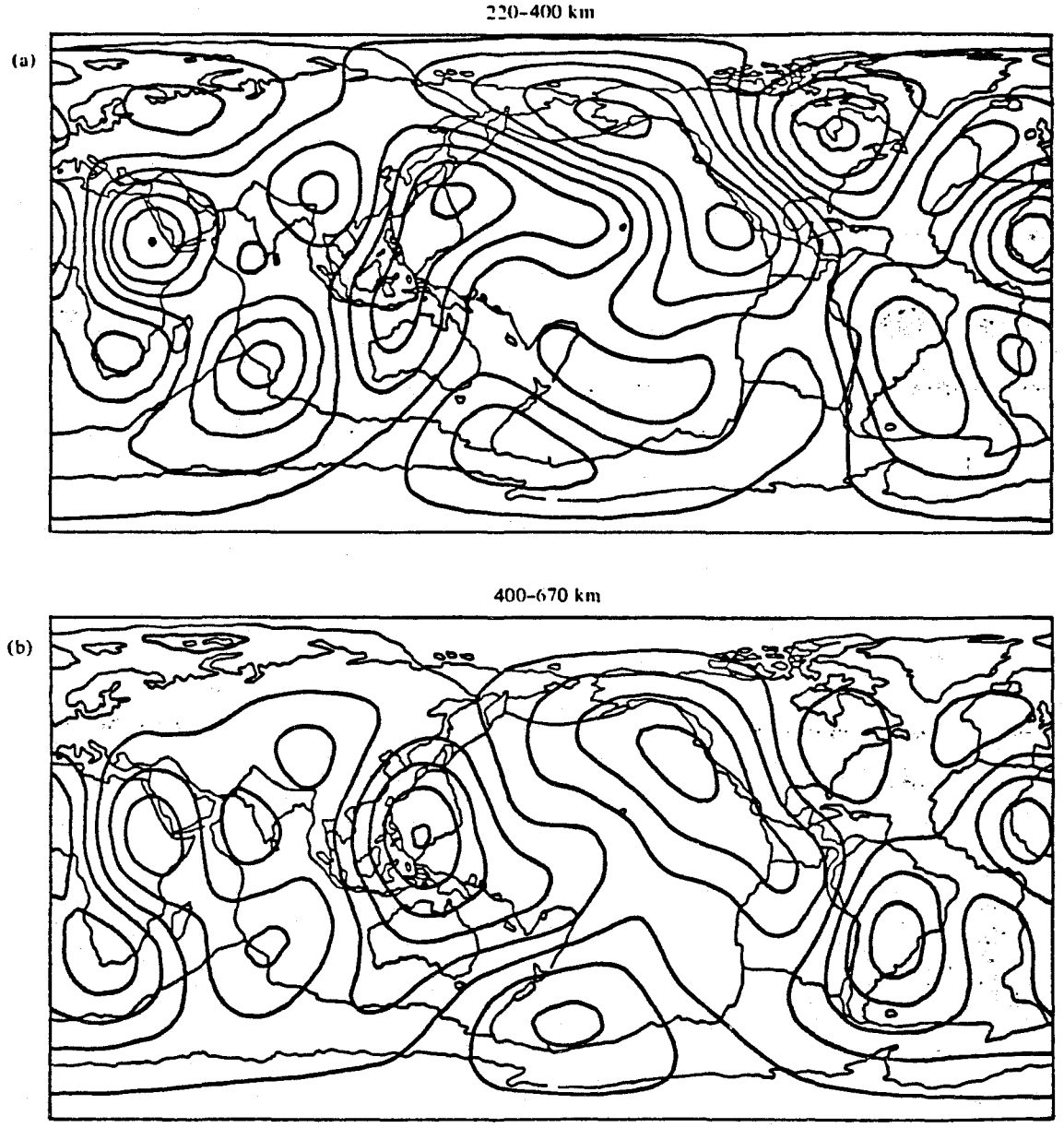


Figure 3.11 Tomographic maps. Global distribution of shear velocity as a function of depth (a–d) as determined by Toshiro Tanimoto of Caltech. The contour interval is 0.5 per cent. Dark regions are faster than average. □

ries and older oceanic regions also have fast seismic velocities (Figure 3.11b). Fast velocities between 400 and 650 km generally correlate with the places where oceanic lithosphere has been overridden by continents since the Mesozoic breakup of the supercontinent **Pangaea.**

Slow velocities under tectonic regions and young oceans can generally be traced to deeper than 300 km (Figure 3.12a), sometimes to 400 km (e.g. the eastern Pacific and northeast Africa, Figure 3.12b), although they are not long linear features (e.g. contrast in the three sub-Atlantic ridge zones crossed in Figure 3.12a–c).

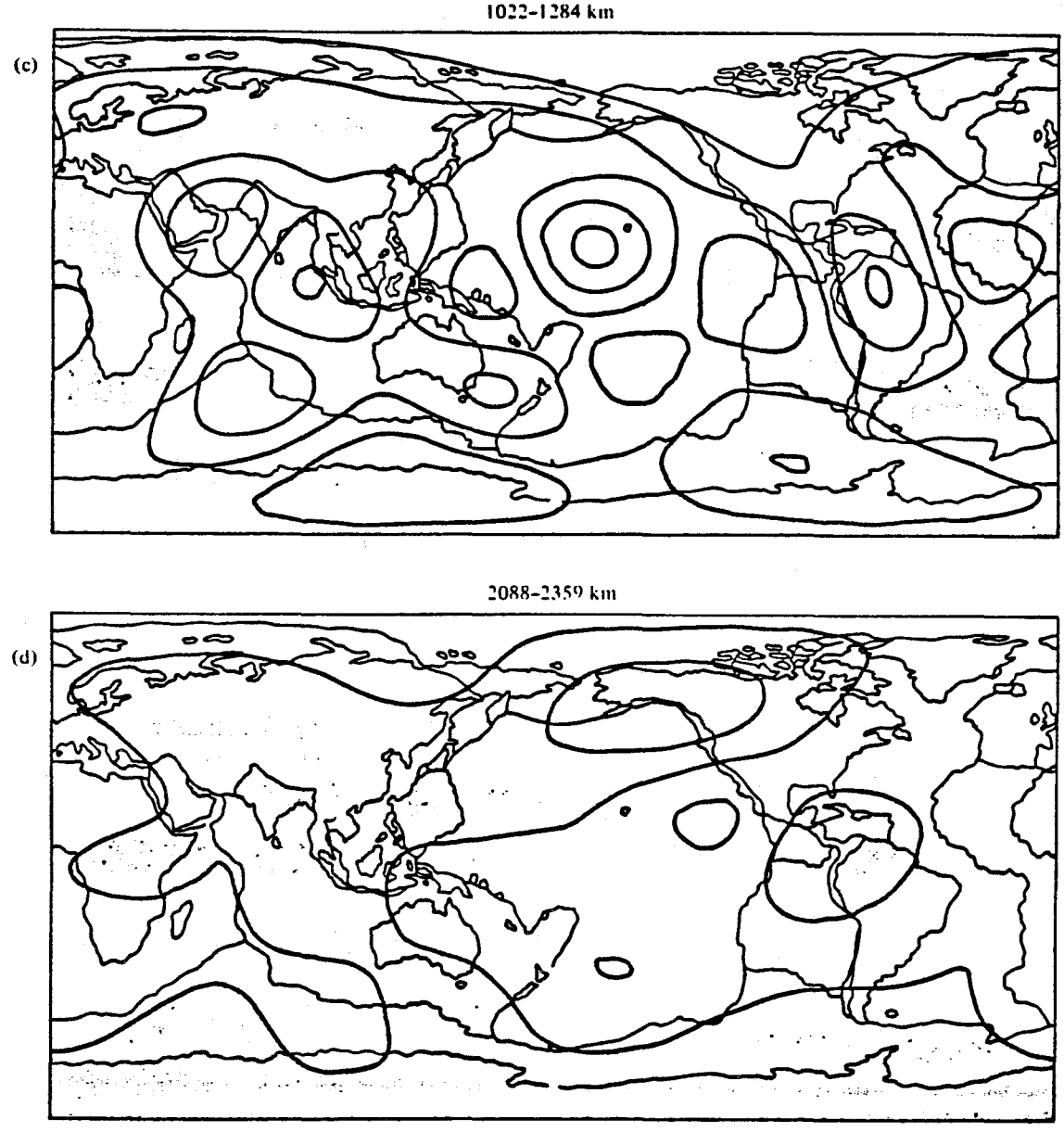
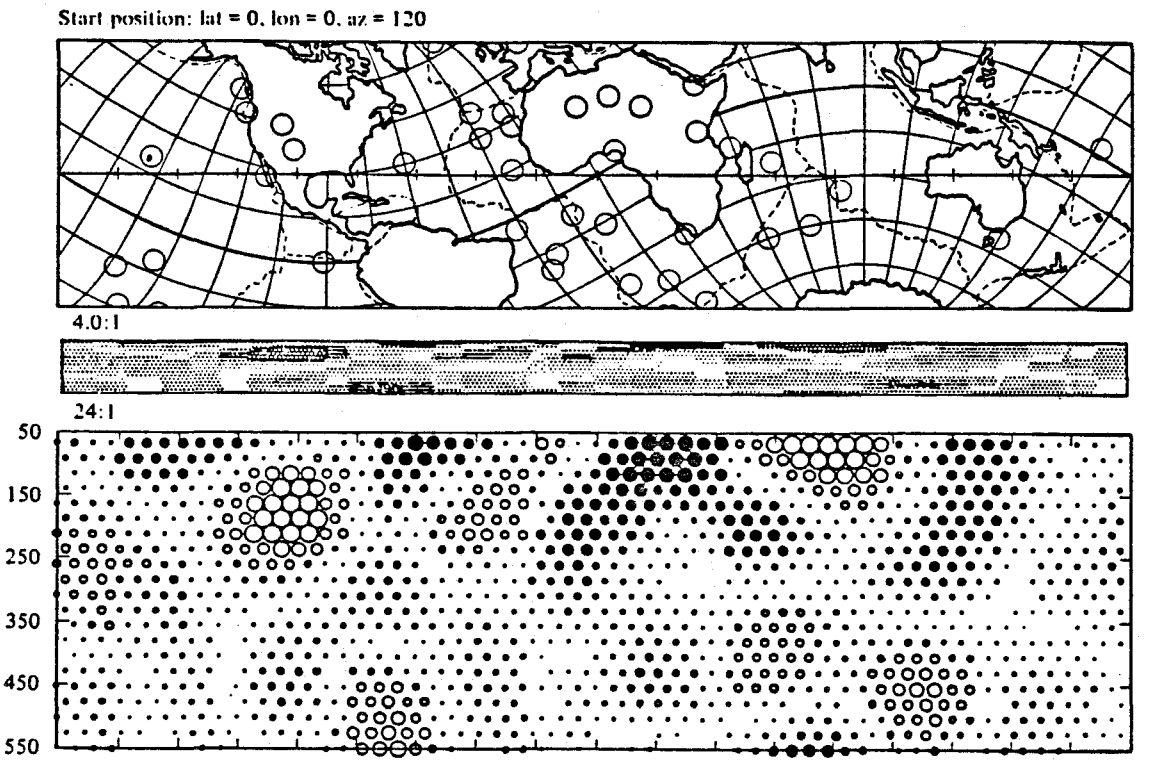


Figure 3.12 Tomographic cross-sections. The shear velocity from 50 to 550 km depth determined from surface waves by H. C. Nataf, I. Nakanishi and Don L. Anderson. The cross-sections are along the great circle path shown in the centre of the upper panel. Plate boundaries (dashes) and hot spots (open circles) are also shown. Cross-sections are given with two different vertical exaggerations (4:1 and 24:1). Open symbols are slower than average velocities; closed symbols are faster than average velocities. □

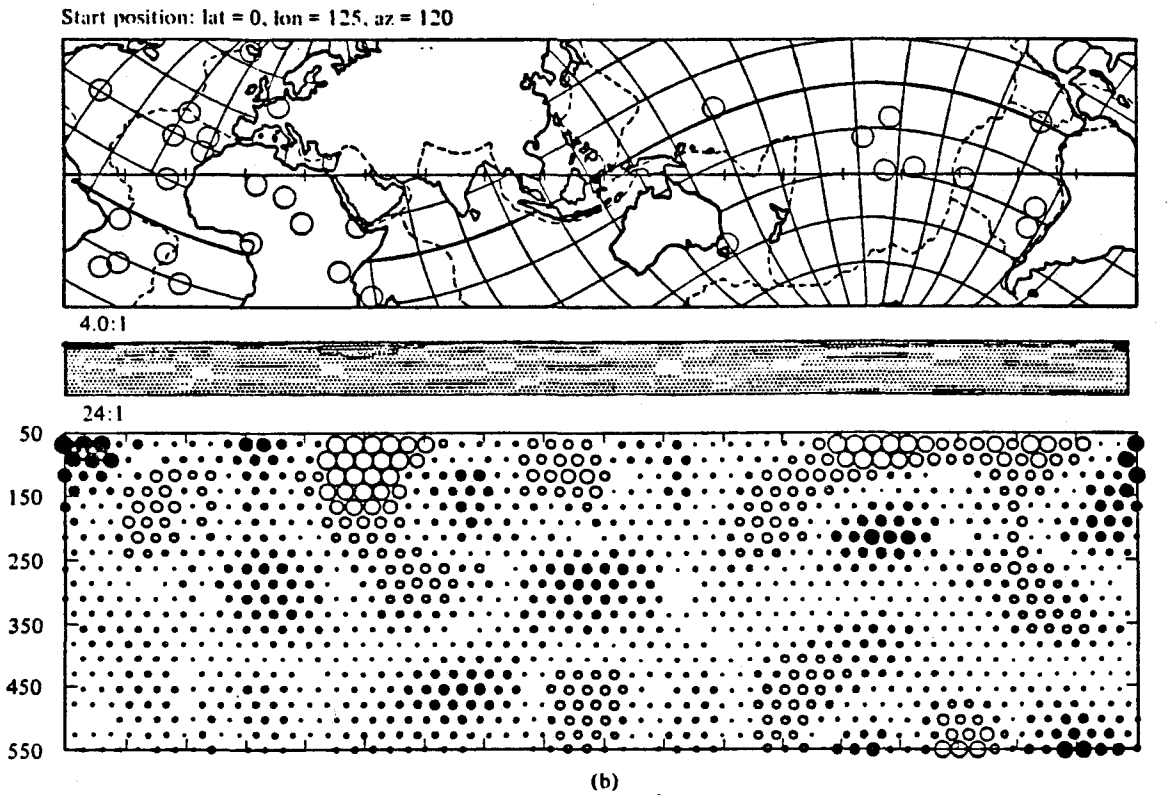


(a)

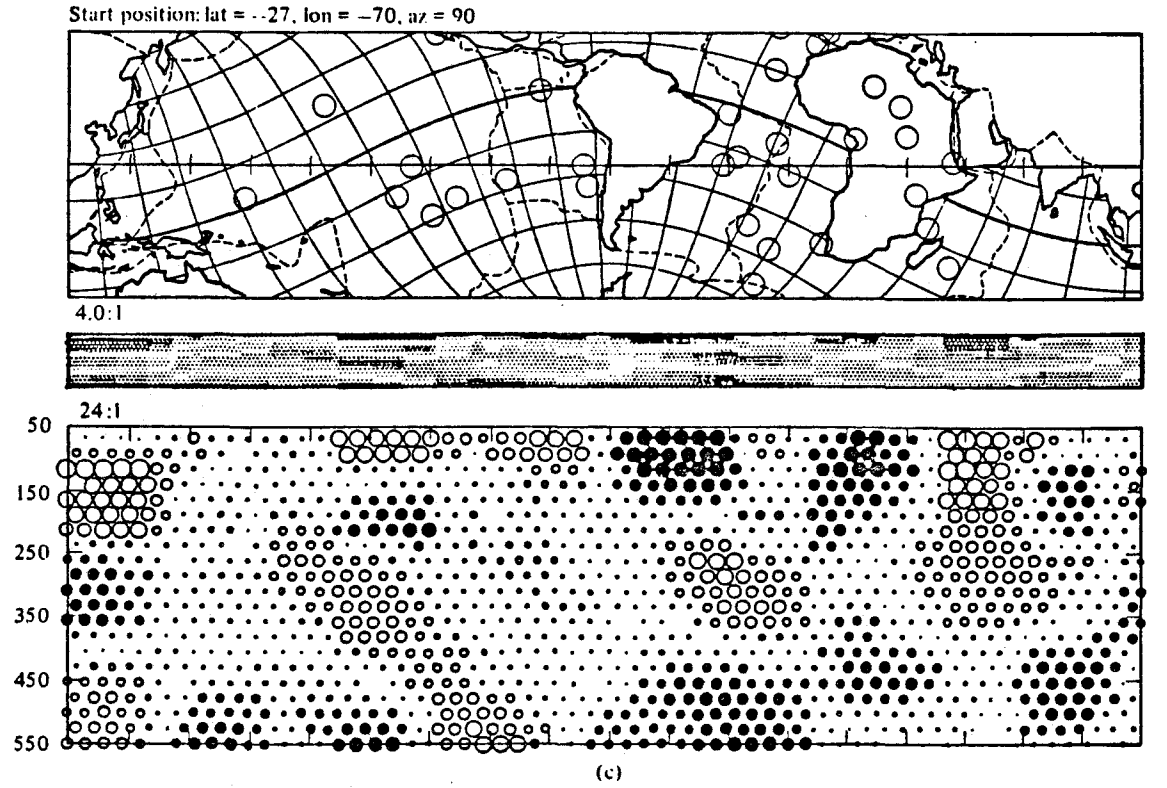
This suggests that the approximate linearity of ocean ridges is due to the fracture properties of the lithosphere rather than to the constancy of the magma source region. Thus, it is not surprising that the deep slow, presumably hot, regions are sometimes offset from the surface expressions of ridges or rifts.

Tomography has fairly poor resolution currently because of the locations and spacings of seismic stations and earthquakes. Features having dimensions of several thousand kilometres can be resolved with global surface wave tomography and, in some regions, body wave tomography can resolve features of a few hundred kilometres in size. With this qualification in mind one gets the impression that mantle upwellings are of large dimension. Upper mantle velocities are very low for large distances away from spreading centres, not just the hundred or so kilometres usually envisaged in cartoons of ocean ridge magmatism, and this is particularly noticeable where the Indian and Pacific ridges are crossed by the Figure 3.12 sections. Oceanic ridges and continental extension regions therefore appear to be embedded in very large regions underlain by seismically slow and, probably, upwelling material. Downwelling slabs are thought to be no thicker than about 100 km and these are not resolved tomographically in the upper 300 km of the mantle. However, below 300 km the shear velocity near subduction zones is higher (e.g. Figure 3.11b near southeast Asia and Figure 3.12b under New Guinea and South America). This suggests that large amounts of cold material are present in the upper mantle in these regions.

Seismic tomography also shows that there are large-scale, low-amplitude lateral variations of seismic velocity in the lower mantle. The lower velocity regions are centred under Africa and its antipode, the central eastern Pacific (Figure 3.11c and d). The low velocities are probably due to high temperatures and these are therefore thought to be upwelling regions of the lower mantle. High velocities occur in a band surrounding the Pacific, including the Pacific rim continents. Most of the continents (except Africa) and most of the subduction zones (except Tonga-Fiji) are above the long-wavelength high-velocity regions of the lower mantle. This may be because continental drift has brought the continents and the associated subduction zones towards downwelling parts of the mantle. If so, prior to the breakup of Pangaea, the supercontinent may have been sitting on top of a hot zone of mantle upwelling.


Start position: lat = 0, lon = 125, az = 120
4.0:1
24:1
50
150
250
350
450
550

(b)


Start position: lat = --27, lon = −70, az = 90
4.0:1
24:1
50
150
250
350
450
550

(c)

The slower than average, upwelling parts of the lower mantle, centred under the central Pacific and Africa, also correlate with highs in the geoid. The geoid is literally 'the shape of the Earth'. Technically, it is the gravitational equipotential surface of the Earth, like the surface of the ocean. The surface of the water is high in the vicinity of excess mass in the crust and mantle. Bumps on the core–mantle boundary also cause highs in the geoid, or equipotential surface. Regions of the mantle which have slow seismic velocity are probably hotter than average and therefore low density and buoyant. This ordinarily would reflect a mass deficiency and would cause a geoid low. However, in dynamic Earth, the buoyant upwelling causes the surface of the Earth and the surface of the core to rise. This elevation of dense material causes the equipotential surface to rise and a geoid high develops. This correlation between tomography and gravity adds confidence to the interpretation of the deep slow zones as regions of upwelling. Hot-spots are generally above these slower regions of the deep mantle. This does not necessarily mean that hot-spots represent material coming from the lower mantle since many hotspots are underlain by faster than average, presumably colder than average, shallow mantle. Even if the mantle convects as a layered system there will be coupling between the upper and lower mantles, either thermal or mechanical. Temperatures in the lower mantle do seem to affect temperatures in the upper mantle, and may therefore have some influence on plate tectonics even if the lower mantle is chemically isolated from the upper mantle.

Conclusions

While we have come a long way from the initial resolution of seismic discontinuities to define the Earth's layered internal structure, much more research is needed if the new results of seismic tomography are to be fully interpreted in terms of the Earth's deep internal dynamics. However, we do have a good picture of the broad compositional state of the Earth's layers and it is now the fine detail of this 'static' view of its interior that needs to be refined with further study; meanwhile, the results from the Earth will provide important lessons if and when future exploration of the planets permits the acquisition there of similarly high quality seismic data.

Further reading

Anderson, D. L. (1989) *Theory of the Earth*, Blackwell Scientific Publications, London, 366 pp.

Bolt, B. A. (1982) *Inside the Earth; Evidence from Earthquakes*, W. H. Freeman, San Francisco, 191 pp.

Brown, G. C. & Mussett, A. E. (1981) *The Inaccessible Earth*, Unwin Hyman, 235 pp. (2nd edition, Chapman & Hall, in press.)